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Ye. P. Novosel'tsev

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ATTENUATION OF TOTAL RADIATION IN UPPER LAYER CLOUDS

Ye. P. Novosel'tsev

ABSTRACT

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The article considers the attenuation of total radiation of ice clouds in the upper layers for different turbidities of the atmosphere below the clouds, and for different values of the albedo of the base surface. The instantaneous and 24-hour average attenuation coefficients are determined for the attenuation of total radiation produced by the clouds of the upper layer.

Author

Until now the question concerning the transmission coefficient of total radiation produced by ice clouds of the upper layer has not been subjected to theoretical considerations. Apparently this is due to the fact that there are no experimental data on a whole series of optical characteristics, which are necessary for the theoretical solution of this problem. Such characteristics include the scattering indicatrix, attenuation coefficient, liquid water content of clouds, etc.

However, it is possible to solve the problem without utilizing these quantities.

According to the results (ref. 1), the magnitude of total radiation at the surface of the earth may be represented in the form

$$Q = \frac{CS_0 \sin h_{\odot}}{e^{\frac{p}{m_{\odot}}\left(1 - e^{m_{\odot}\tau^*}\right)} + (1 - a_c)\,\frac{2}{m_{\odot}}\,\Gamma_D\, e^{-\frac{p}{m_{\odot}} e^{m_{\odot}\tau^*}}\, F}, \qquad (1)$$

where $F \equiv \left[Ei\left(\frac{p}{m_{\odot}}\right) - Ei\left(\frac{p}{m_{\odot}} e^{m_{\odot}\tau^*}\right)\right]$; C is the transmission coefficient of total radiation of the atmosphere below the clouds; S_0 is the solar constant; $h_{\odot}$ is the altitude of the sun above the horizon; $m_{\odot}=\frac{1}{\sin h_{\odot}}$; a_c is the magnitude of the albedo of the system consisting of the atmosphere below the clouds and the earth; τ^* is the optical thickness of clouds; $p=(2\Gamma_D - m_{\odot}\beta)e^{-m_{\odot}\tau^*}$; $\beta_{\odot}$ is some function of the scattering indicatrix which characterizes the portion of direct radiation scattered into the upper hemisphere; Γ_D is also a function of the scattering indicatrix which determines the portion of diffusion radiation scattered into the upper hemisphere (ref. 2), and Ei(x) is an integral exponential function. Coefficients c and a_c may be computed quite reliably, using the method suggested in refs. 3, 4, and 5,

$$C=\frac{Q_1}{m^*(1+fm^*)}, \quad a_c=\frac{a_s+fm^*}{1+fm^*},$$

where Q_1 is the magnitude of the total radiation at the lower boundary of the clouds, and m* is the average secant of the angle which determines the direction of distribution of the center of gravity of the descending radiation at the level of the lower boundary of the clouds. According to the data in reference 1, $m^*=(m_{\odot}-2)e^{-m_{\odot}\tau^*}+2$, f is a coefficient which characterizes the transparency of the atmosphere below the clouds (ref. 5), and $a_{surface}$ is the albedo of the base surface.

To compute coefficients Γ_D and $\beta_{\odot}$, it is necessary to know the scattering of the ice particles.

It turns out that the indicatrix for the scattering of radiation by cloud particles may be determined, approximately, by proceeding from simple considerations.

As we know, the clouds of the upper layer consist of small crystals which can unite and form nodules consisting of a large number of needles and plates. Due to multiple reflection from the boundaries of such formations, these nodules will scatter radiation by diffusion. In addition, the ice crystals are ingrained with air, due to which their scattering power is quite high.

In view of what we have said above, we may consider the clouds of the upper layer, in the first approximation, as a medium consisting of particles which scatter radiation uniformly in all directions, with the exception of a very small angle in the direction of propagation of the incident radiation.

Any indicatrix for sufficiently large (compared with wavelength) particles may be obtained by totaling the effects of reflection and diffraction. The error produced by the fact that we add intensities, and not the fields, does not exceed several percent.

For this reason the scattering indicatrix $x_1(\gamma)$, due to reflection, apparently has a form close to spherical. The indicatrix due to diffraction represents a long, narrow "nose." In this case all of the diffracted energy is concentrated within the limits of a small angle. Thus, the total indicatrix may be represented in the following form:

$$x(\gamma) = p_1 x_1(\gamma) + p_2 x_2(\gamma),$$

where p_1 and p_2 are the relative "weights" of the magnitudes of reflected and diffracted radiation. As we know, for large particles,

$$p_1 = p_2 = \frac{1}{2}.$$

Thus,

$$x(\gamma) = \frac{1}{2}[1 + x_2(\gamma)],$$

because for a spherical indicatrix $x(\gamma) = 1$.

According to reference 2,

$$\beta(h_\odot) = \frac{1}{4\pi}\int_0^{2\pi} d\varphi \int_{\pi/2}^{\pi} x(\gamma)\sin\vartheta \, d\vartheta,$$

where $$\gamma = \arccos(\cos\vartheta \sin h_\odot + \sin\vartheta \cos h_\odot \cos\varphi),$$

and

$$\Gamma_D = \frac{1}{2}\int_0^{2\pi} d\varphi \int_{\pi/2}^{\pi} \beta(\vartheta)\sin\vartheta\, d\vartheta.$$

If we note that in practice the indicatrix $x_2(\gamma)$ differs from zero only within a narrow cone, the quantities $\beta_\odot$ and $\Gamma_\odot$ may be determined immediately: $\beta_\odot = \Gamma_D = 0.25$.

The last parameter which must be determined is the optical thickness of the upper cloud layer.

The average magnitude of the thickness of the upper cloud layer was determined by us on the basis of averaging data collected over many years during measurement of direct solar radiation, which has passed through the upper cloud layer as well as that which was attenuated only by the cloudless atmosphere.

Let $I_1(h_\odot)$ be the average value of the direct radiation, when the altitude of the sun is $h_\odot$ and when the sky is clear. Let $I_2(h_\odot)$ be the average magnitude of the direct radiation which has passed through the upper cloud layer, when the altitude of the sun was the same; let τ_1 be the average optical thickness of the cloudless atmosphere, τ_2 be the average optical thickness of the cloud layer, and I_0 be the solar constant.

Then

$$I_1 = I_0 e^{-\frac{\tau_1}{\sin h_\odot}} \text{ and } I_2 = I_0 e^{-\frac{(\tau_1+\tau_2)}{\sin h_\odot}},$$

from which it follows that

$$\frac{I_2}{I_1} = e^{-\frac{\tau_2}{\sin h_\odot}}$$

or

$$\tau_2 = -\sin h_\odot \ln\frac{I_2}{I_1}. \qquad (1)$$

By using equation (1) we computed the average values of τ_2 for various altitudes of the sun.

However, τ_2 is still not the true optical thickness of the cloud layer. The fact is that, as we have already stated, all the diffracted energy is concentrated inside of a small solid angle, and therefore this part of the scattered energy (it constitutes exactly half of the entire scattered radiation) enters the input aperture of the actinometer. Consequently, the optical thickness determined from equation (1) will be less than the true value by a factor of τ^*.

In view of the fact that our problem consists of determining the attenuation coefficient for the upper layer clouds as a whole, we shall not separate the clouds of this formation into types Ci, Cs, Cc. Therefore, Table 1 shows the average optical thickness of clouds Ci and Cs obtained by means of the following equation:

$$\tau^* = \frac{\tau^*_{Ci} + \tau^*}{2}.$$

TABLE 1

$h^\circ_\odot$ . .	15	20	25	30	40	50
τ^* . . .	1.6	1.35	1.00	0.90	0.63	0.47

As we can see from the table, the magnitude of the optical thickness of the upper cloud layer has a noticeable diurnal variation. This is explained by the fact that as the altitude of the sun above the horizon increases and is accompanied by an increase in the flux of solar radiation, the magnitude of the absorbed energy also increases, so that the clouds are partially evaporated.

Now we have all of the data necessary to compute the total radiation when we have an upper cloud layer.

Figure 1 shows the variation in total radiation in the upper cloud layer as a function of the altitude of the sun above the horizon, which we have computed. The calculations were made for two states of the atmosphere below the clouds: for an atmosphere with high transparency characterized by the parameter $f = 0.10$, and with a low transparency ($f = 0.15$). The albedo of the base surface was assumed to be equal to 0.2. The same figure shows the curves corresponding to the magnitudes of total radiation when the atmosphere is without clouds and for the same values of parameters f ($f_1 = 0.10$, $f_2 = 0.15$).

Figure 2 shows the variation in the attenuation of total radiation by the clouds of the upper layer $C_{upper}(h_\odot)$ as a function of solar altitude above the

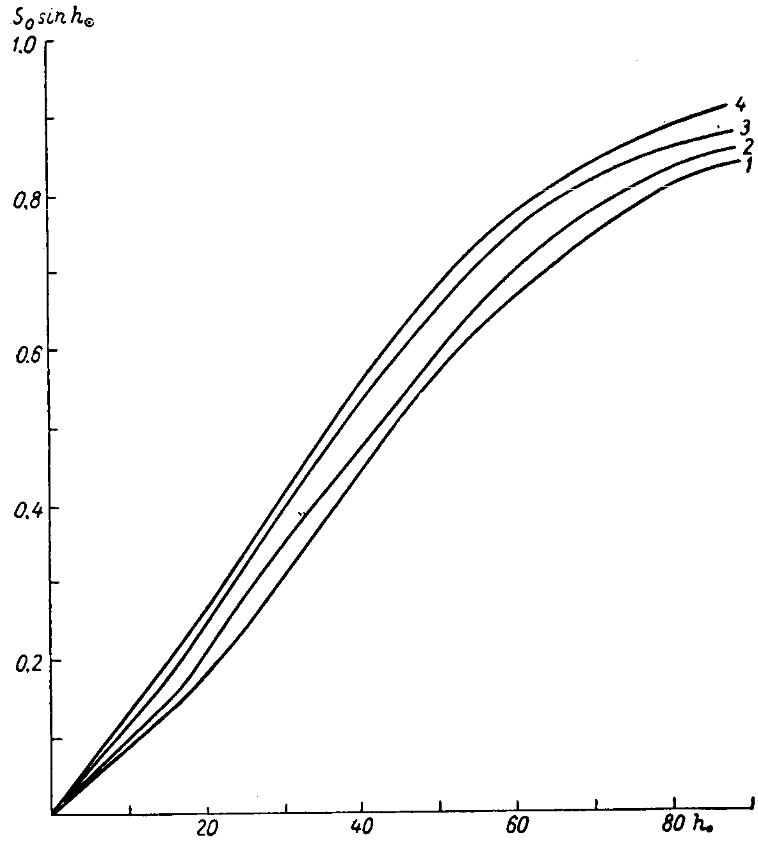

Figure 1. Magnitude of total radiation for different altitudes of sun above horizon. For clouds of upper formation: 1, f = 0.14; 2, f = 0.10; for clear sky: 3, f = 0.14; 4, f = 0.10.

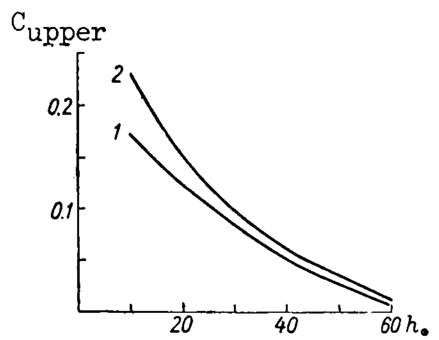

Figure 2. Variation in attenuation coefficient C_{upper} as function of altitude of sun above horizon. 1, a_e = 0.65; 2, a_e = 0.20.

horizon. By examining the figure, we can see that the magnitude of the coefficient $C_{upper}(h_{\odot})$ depends substantially upon the altitude of the sun. The relationship obtained correlates sufficiently with experimental data. The sharp variation in C_{upper} as a function of solar altitude makes it difficult to use this coefficient for the solution of a series of meteorological problems, particularly since in the solution of many problems we cannot use the instantaneous value, but must have a value averaged over a period of one day, ten days, one month, etc.

In connection with this we computed the diurnal attenuation coefficients $\overline{C}_{upper}$. This quantity is obtained from the following equation:

$$\overline{C}_{upper} = 1 - \frac{\Sigma_2}{\Sigma_1},$$

where Σ_1 is the diurnal sum of the total radiation for a clear sky, Σ_2 is the total diurnal radiation when we have an upper cloud layer. The diurnal sums of total radiation for a clear sky may be obtained quite simply (refs. 5 and 6).

It is difficult to obtain an analytic expression for the diurnal sum of the total radiation when we have clouds in the upper layer, because it is necessary to integrate expression (1) over the time from sunrise to sunset. Therefore, we carried out numerical integration in the manner described. The daylight period was broken down into hourly intervals, and for each interval the average altitude of the sun was determined. From this altitude of the sun, by using figure 1, the corresponding value of the total radiation was determined. By multiplying this quantity by 60, we obtained the hourly sums. As a result of the summation of all hourly sums from sunrise to sunset, we determined the diurnal sums of the total radiation.

The average diurnal coefficients C wer computed by us for all months and for various latitudes (from 30-70°). As was to be expected, these coefficients $\overline{C}$ have a substantial annual variation and depend noticeably on the latitude of the location. These relationships are shown in figures 3 and 4. Although there is a noticeable variation in the magnitude of $\overline{C}$ as a function of the time of year and the latitude of the location, the utilization of this coefficient for the large number of meteorological problems is substantially more convenient than the utilization of coefficient $C(h_{\odot})$, because this coefficient varies over a smaller range. Indeed, for each latitude we may select two to three seasons during which the variation in $\overline{C}$ may be neglected. In addition to the variation in the attenuation coefficient as a function of these factors, we must also consider its variation with the albedo of the base surface.

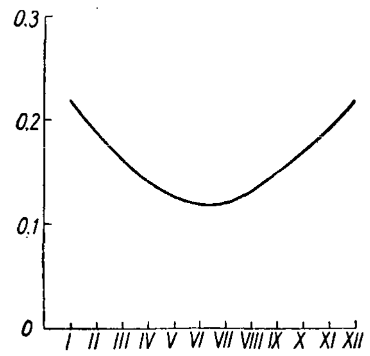

Figure 3. Annual variation in average diurnal attenuation coefficient.

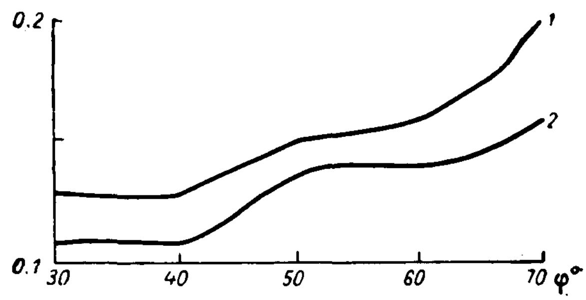

Figure 4. Variation in attenuation coefficient with latitude (June). 1, f = 0.14; 2, f = 0.10.

As illustration, figure 2 shows the diurnal variation of the attenuation coefficient for two values of the albedo of the base surface $a_e = 0.2$ and $a_e = 0.65$ ($a_e = a_{earth}$). We can see that when all other conditions are equal, the attenuation for $a_e = 0.65$ is substantially less than for $a_e = 0.20$.

A similar conclusion should be made concerning the variation in the average diurnal attenuation coefficient $\overline{C}$ as a function of the magnitude of the albedo of the base surface.

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